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16. ABSTRACT

The characteristic features of the turbulent atmospheric boundary layer (ABL) are reviewed by means of scaling arguments. The goal was to characterize the meteorological conditions affecting diffusion for transport on the order of 10 km or less. The idealized ABL is divided into a number of regions, each characterized by distinct scaling parameters for the turbulence. With these scaling parameters wind and vertical dispersion are parameterized. The lateral dispersion is treated separately. Methods are suggested for estimating the surface fluxes of heat and momentum, and the mixing height. These are needed for the characterization of diffusion within each of the regions of the ABL. The diffusion is often found to be other than Gaussian in the vertical direction. Several of the models characterize directly the crosswind integrated concentration at the surface, thus avoiding whenever justified the assumption of a Gaussian distribution of material in the vertical. Comparisons are presented for the surface releases conducted during Project Prairie and the nonbuoyant elevated releases conducted at Copenhagen.

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1985

PARAMETERIZATION OF THE ATMOSPHERIC BOUNDARY LAYER FOR AIR POLLUTION DISPERSION MODELS

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INTRODUCTION

Most of the gaseous pollutants that are emitted into the air are released into the Atmospheric Boundary Layer (ABL). The main characteristic of the ABL is turbulence. Turbulent flows have a wide range of length scales. The scales are bounded from above by the geometry of the flow field and from below by molecular effects. The structure of turbulence is determined by the various sources as well as by the imposed length and time scales. The turbulence in the ABL, therefore, is multifacet. The outstanding characteristic of turbulence is its immense ability to disperse and transport momentum, heat and also contaminants. Therefore the laws, that characterize the turbulence also characterize the atmospheric dispersion (Tennekes and Lumley, 1972; Pasquill and Smith, 1983).

The complex nature of turbulence does not allow that a single model is able to describe the dispersion process under all the apparent forms of atmospheric turbulence. Nevertheless, most of the estimates of dispersion from point sources were, and often still are, based on the Gaussian plume model. Throughout the years a multitude of schemes have emerged, which describe the input parameters of this model (Sutton, 1953; Sedifian and Bennet, 1980). Often, the input parameters are related to simple turbulence typing schemes or stability classes. A problem with these stability classes is that they cover a broad range of stability conditions and that they are very site specific. Moreover, these classes are biased toward neutral stability when unstable or convective conditions in the ABL actually exist (Weil, 1983; Berkowicz et al., 1983). The influence of these aspects on the calculated ground level concentrations is large (Kretzschmar and Mertens, 1984).

Over the past decade considerable progress has been made in our understanding of the turbulent structure of the ABL (Caughey, 1982; Tennekes, 1982; Wyngaard, 1982; Nieuwstadt; 1984a; Panofsky and Dutton, 1984). This allows us to describe the dispersion process more closely to the turbulent structure than is used in the traditional plume models (Hanna et al., 1977; Willis and Deardorff, 1978; Venkatram et al., 1984). In this paper we review the characteristic scaling parameters and the structure of the turbulence. The idealized ABL is divided into a number of regions, each characterized by distinct scaling parameters for the turbulence. With these scaling parameters wind and vertical dispersion can be parameterized. Lateral dispersion is treated separately.

For several scaling regions we suggest models for the dispersion in the vertical direction, the choice of which depends on the actual state of the atmosphere. The models directly give the cross-wind integrated concentrations at the ground χ_y for non-buoyant releases of continuous point sources. Whenever justified, the assumption of Gaussian dispersion in the vertical is avoided. In contrast the lateral concentration distribution is taken as Gaussian. The parameters in this distribution are briefly summarized. With the lateral distribution and χ_y we can calculate the concentrations at the surface for a given position. This will be demonstrated in this paper.

In the following discussion we limit ourselves to horizontal homogeneous conditions over land and a travel distance less than ≈ 10 km. Often the distance to the maximum ground level concentrations is within this range. The corresponding travel time for the pollutants is roughly 30 min to 1 h. We begin the discussion with a brief description of the idealized ABL over land. The ideas are illustrated with meteorological data from the Cabauw mast (Driedonks et al., 1978). The suggested dispersion models are compared with data from the Prairie grass experiment for near surface releases (Barad, 1958) and data from Copenhagen (Gryning and Lyck, 1984) for elevated releases. An accompanying report (Irwin et al., 1985) provides more details.

GENERAL CHARACTERISTICS OF THE ABL OVER LAND

The ABL can be defined as the lower layer of the atmosphere, where the influence of the surface is present due to friction and due to heating and cooling. Friction between the flow and the surface causes the shear production of mechanical turbulence in the ABL. This results in a flux of momentum to the surface τ_o. Heating of the surface may cause the production of thermal turbulence. This is the case during daytime when the surface is heated by solar radiation, which normally results in a upward sensible heat flux H. This leads to an unstable state of the ABL ($H > 0$). During nighttime the surface usually cools by longwave emission of radiation. In that case the turbulence of mechanical origin is suppressed by a downward heat flux H, which results in a stable stratification of the ABL ($H < 0$). Therefore, both during daytime and nighttime H puts constraints on the ~~existence~~ existing level of turbulence.

The sensible heat flux H is related to the net radiation Q^*, which is the driving force of the surface energy budget

$$H + LE + G = Q^*. \qquad (1)$$

Here LE is the flux of energy due to evaporation ($LE > 0$) or condensation ($LE < 0$) and G is the soil heat flux. A typical example of the observed diurnal variation of H, LE, G and Q^* in summertime at clear skies is given

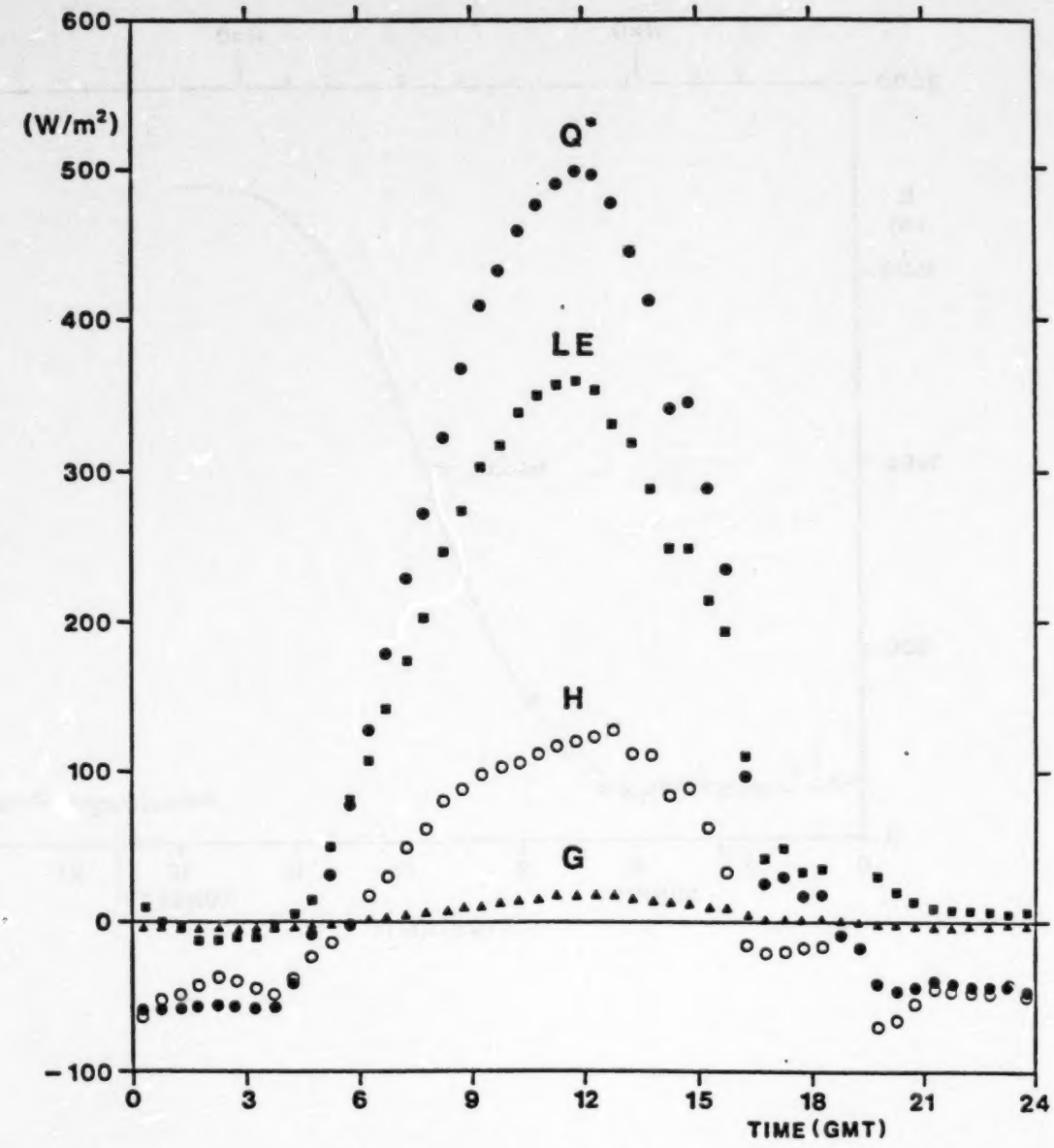

Fig. 1 The observed diurnal variation of the components in the surface energy budget of Eq. (1), at Cabauw on a cloudless day in summertime (May 31, 1978). Here
Q^* is net radiation,
LE is flux due to evaporation,
H is sensible heat flux, and
G is soil heat flux.

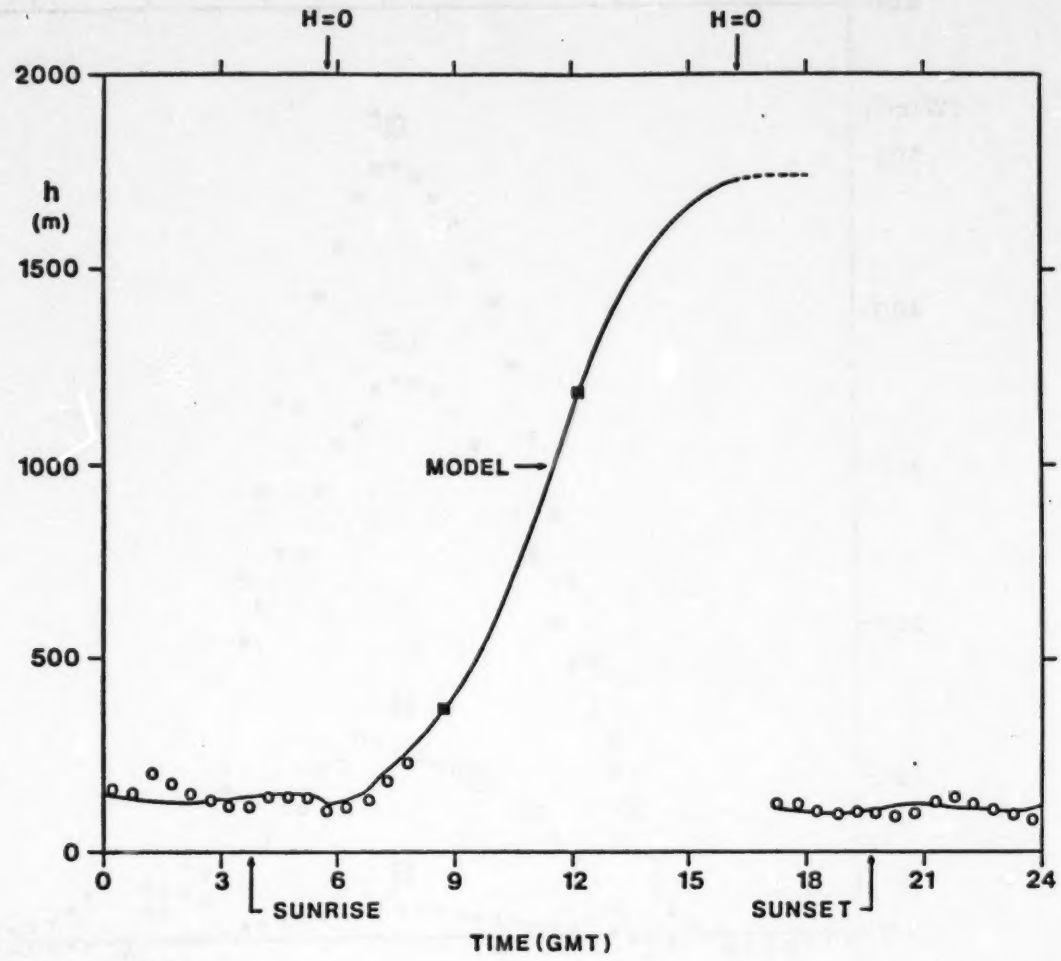

Fig. 2 The diurnal variation of mixing height h for the period in Fig. 1. Indicated are the moments for sunrise, sunset and H = 0. Dots indicate observations of h with an acoustic sounder, squares are estimates of h obtained from temperature profiles. Further explanation is in the text.

in Fig. 1. The data of Fig. 1 is discussed in De Bruin and Holtslag (1982). During daytime Fig. 1 shows that H, LE and G are roughly proportional to Q^*, while G is relatively small. The partitioning of H and LE over the available energy $Q^* - G$ depends primarily on the surface moisture condition (Monteith, 1981; De Bruin and Holtslag, 1982). Holtslag and Van Ulden (1983) discuss a simple parameterization scheme for the terms in (1) during daytime using routine weather data only (e.g. total cloud cover, air temperature at one height). During nighttime the behaviour of the terms in (1) is more complicated. In that case both H and Q^* are strongly influenced by wind speed and by temperature differences near the surface (Van Ulden and Holtslag, 1983).

The surface fluxes of heat and momentum H and τ_o respectively, define a length scale L as (Monin and Yaglom, 1971)

$$L = - \frac{u_*^3}{k \frac{g}{T} \overline{w\theta}_o}, \qquad (2)$$

for dry air. Here L is known as the Obukhov length, u_* is called the friction velocity, k the Von Kármán constant ($k \approx 0.4$), g/T the buoyancy parameter and $\overline{w\theta}_o$ the kinematic surface heat flux. The latter flux is related to H by $H = \rho c_p \overline{w\theta}_o$. The surface momentum flux τ_o defines u_* by $\tau_o = u_*^2$ (kinematic units). Near the surface the height z and |L| define the relative importance of shear production and thermal production (or dissipation) of turbulence. In the unstable ABL the thermal production is larger above $-z/L \approx 0.6$, while below the latter height the shear production is larger (Businger, 1973). As such |z/L| is an important parameter for the structure of the turbulence in the surface layer. This is the basis of the Monin-Obukhov similarity theory (Monin and Yaglom, 1971; Tennekes, 1982).

Above the surface layer the structure of the turbulence is influenced by the mixing height h as well. This height is defined as the height to which turbulence extends and to which pollutants are generally well-mixed. In daytime h roughly coincides with the height of the ABL, typically to the base of the first inversion. During nighttime h is typically smaller than the height up to which significant cooling extends (Wetzel, 1982). Fig. 2 shows the diurnal variation of the mixing height h for the period in Fig. 1. During nighttime h was observed with an acoustic sounder (indicated with dots) (Nieuwstadt, 1984b). An acoustic sounder measures the height of the layer in which temperature fluctations exist. This height coincides with h because the temperature fluctuations are generally caused by turbulence. The indicated line is during nighttime obtained with a model to be discussed below. During daytime h is calculated with a so-called mixed-layer model (Driedonks, 1982). This model used the actual measured values of τ_o and H for each half hour and the temperature profile of the morning sounding. The model confirms the observation of h with the sounding at 12 h. In the afternoon transition period the dashed line indicates that the turbulence is not dissipated immediately (Benkley and Schulman, 1979), while at the same time a stable layer is formed near the surface.

From Fig. 2 it is seen that the afternoon value of h is an order of magnitude larger than the nighttime value. This will influence the structure of the turbulence and also the dispersion of air pollution. In the unstable ABL the largest eddies are related to h (Caughey, 1982), while in the stable ABL the eddies are related to local parameters (Nieuwstadt, 1984a). Together with the already introduced length scales z and L we can form two independent dimensionless parameters for the

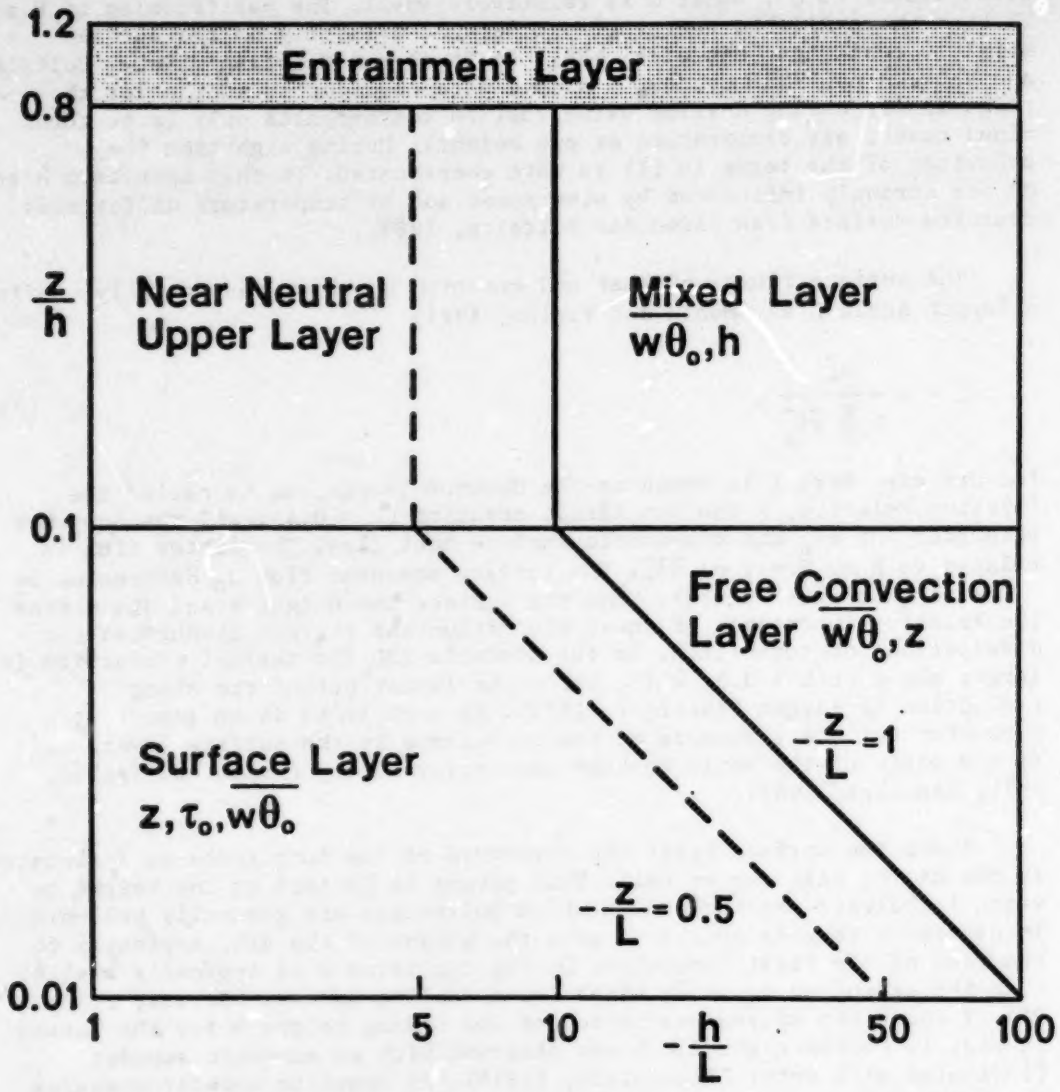

Fig. 3 Definition of scaling regions in the unstable ABL (L < 0) with indicated basic scaling parameters for the turbulence (Holtslag and Nieuwstadt, 1985).

turbulent structure in the ABL. Here we choose z/h, the relative height and h/|L|. The latter ratio can be regarded as a stability parameter for the whole mixing layer. For small h/|L| the stratification is close to neutral, while for increasing h/|L| stability effects become more important in the whole ABL.

Fig. 3 taken from Holtslag and Nieuwstadt (1985) shows a plot of z/h against a typical range of - h/L for the unstable boundary layer. In the diagram five regions are distinguished, which are commonly mentioned in literature for an idealized ABL (Panofsky, 1978; Caughey, 1982). The dividing lines between the regions are discussed in the next sections. The diagram is similar to those presented by Nichols and Readings (1979) and Olesen et al. (1984), but they use z/h and z/L as independent parameters.

The diagram of Fig. 3 identifies the principle scaling parameters for the turbulence in the surface layer, the free convection layer and the mixed layer. The turbulent structure of these layers will be discussed in the next sections. We will not go into detail for the structure of the near neutral upper layer and the entrainment layer. At present the theory for these layers is less well established (Deardorff et al., 1980; Driedonks and Tennekes, 1984). Below also the counterpart of Fig. 3 for the stable ABL is given. With the aid of these figures we discuss vertical dispersion and wind profiles. Lateral dispersion is discussed separately at the end of this paper. (Fig. 10)

THE SURFACE LAYER (SL)

The SL is defined as the lower layer in the ABL in which wind shear plays a dominant role (Businger et al., 1971). Therefore τ_o is an important scaling parameter. Because of surface heating or cooling also $\overline{w\theta}_o$ is of importance. Normally, τ and $\overline{w\theta}$ decrease with height. If we accept a 10% variation of the fluxes, then in the case of linear profiles the SL extends to $z/h \approx 0.1$. In the unstable ABL the flux τ_o is no longer important as a scaling parameter, when the production of turbulent kinetic energy due to buoyancy is comparable to that of the shear stress. This defines the top of the SL roughly as the minimum of $z/h \approx 0.1$ and $|z/L| \approx 1$.

For the SL Monin-Obukhov theory applies, which means that turbulent quantities can be expressed as a function of z/L (Businger et al., 1971). For instance, the non-dimensional wind gradient can be written as

$$\frac{kz}{u_*}\frac{\partial U(z)}{\partial z} = \phi_m(z/L) \ . \qquad (3)$$

Here U(z) is wind speed at height z and ϕ_m is an empirical function of z/L, which for unstable conditions can be written as (Dyer, 1974; Carl et al., 1973)

$$\phi_m = (1 - 16\ z/L)^{-m} \ . \qquad (4a)$$

Dyer proposes m = 1/4, while Carl et al. propose m = 1/3. The latter corresponds to the theoretical limit in free convection (Panofsky, 1978). For stable conditions (Dyer, 1974)

$$\phi_m = 1 + 5\ z/L \ . \qquad (4b)$$

From (3), (4a) and (4b) the wind speed profile in the SL can be

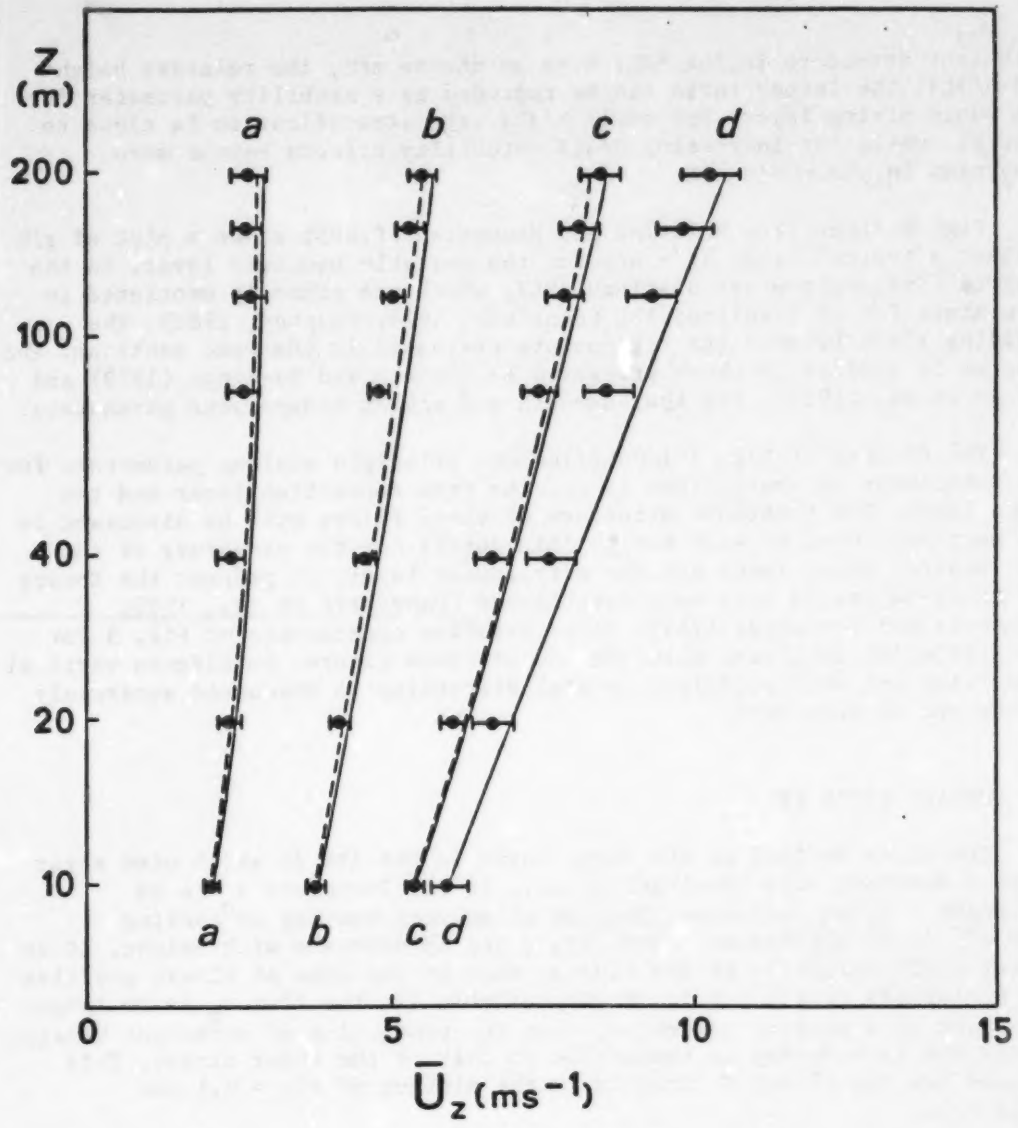

Fig. 4 The variation of mean wind speed $\bar{U}_z$ with height z at Cabauw for 4 stability classes defined as:
a: $-40 \leq L < -12$ m, $L_m = -28$ m
b: $-200 \leq L < -40$ m, $L_m = -95$ m
c: $-1000 \leq L < -200$ m, $L_m = -365$ m
d: $|L| > 1000$ m, $L_m = 10^4$ m.
Here L_m is a central value for L. The solid line is Eq. (5) with ψ_M based on (4a) and m = 1/4, while the dashed line gives the result for m = 1/3. Observations are indicated with dots and error bars (Holtslag, 1984b).

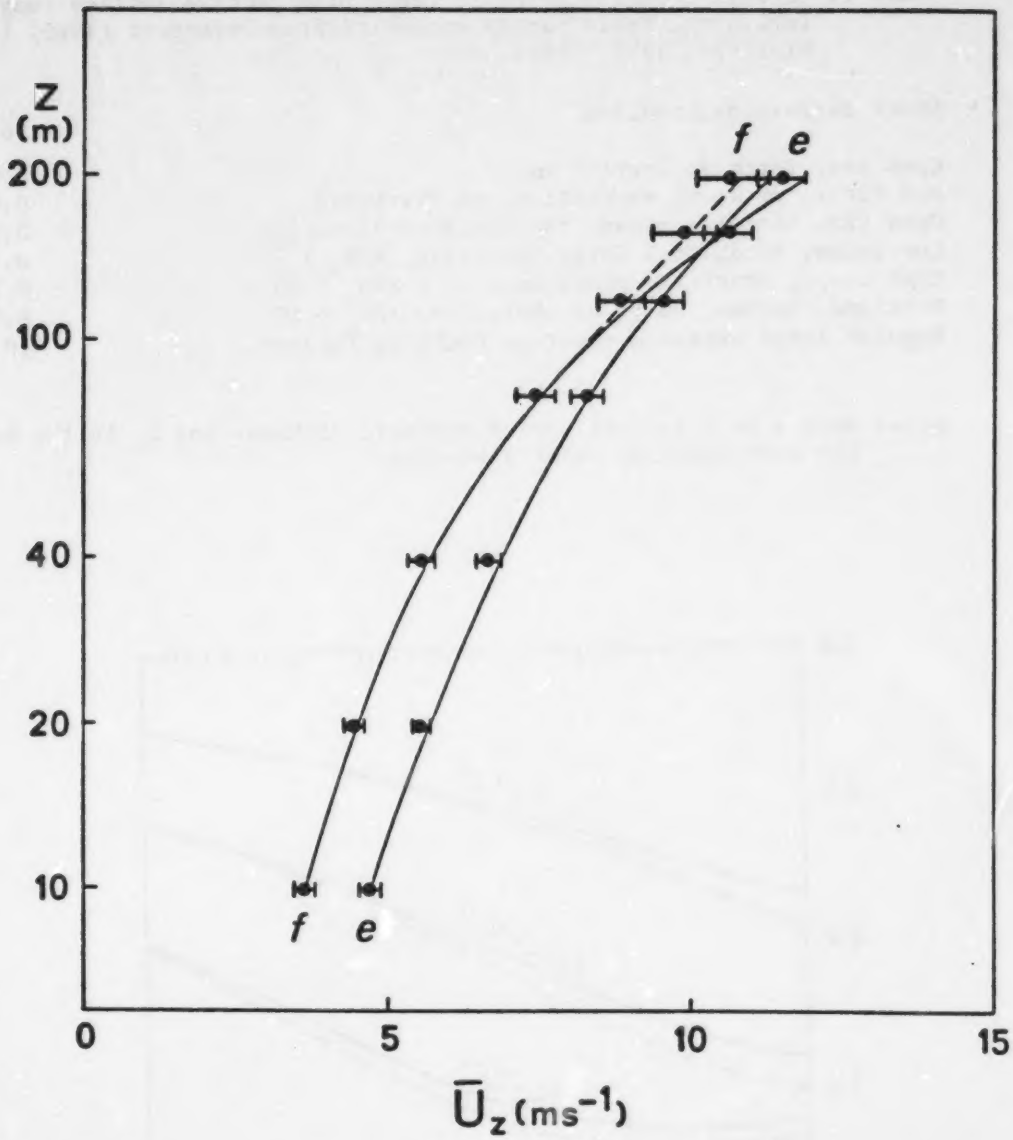

Fig. 5 As Fig. 4 but for two stable classes (Holtslag, 1984b):
e: $200 < L \leq 1000$, $L_m = 350$.
f: $100 < L \leq 200$, $L_m = 130$.

Table 1. Terrain classification in terms of effective surface roughness length z_o. Table partly extracted from Davenport (1960) (See also Wieringa, 1980, 1981).

Short terrain description	z_o(m)
Open sea, fetch at least 5 km	0.0002
Mud flats, snow; no vegetation, no obstacles	0.005
Open flat terrain; grass, few isolated obstacles	0.03
Low crops, occasional large obstacles, $x/h_o > 20$	0.10
High crops, scattered obstacles, $15 < x/h_o < 20$	0.25
Parkland, bushes, numerous obstacles, $x/h_o \sim 10$	0.5
Regular large obstacle coverage (suburb, forest)	(0.5-1.0)

Note: Here x is a typical upwind obstacle distance and h_o is the height of the corresponding major obstacles.

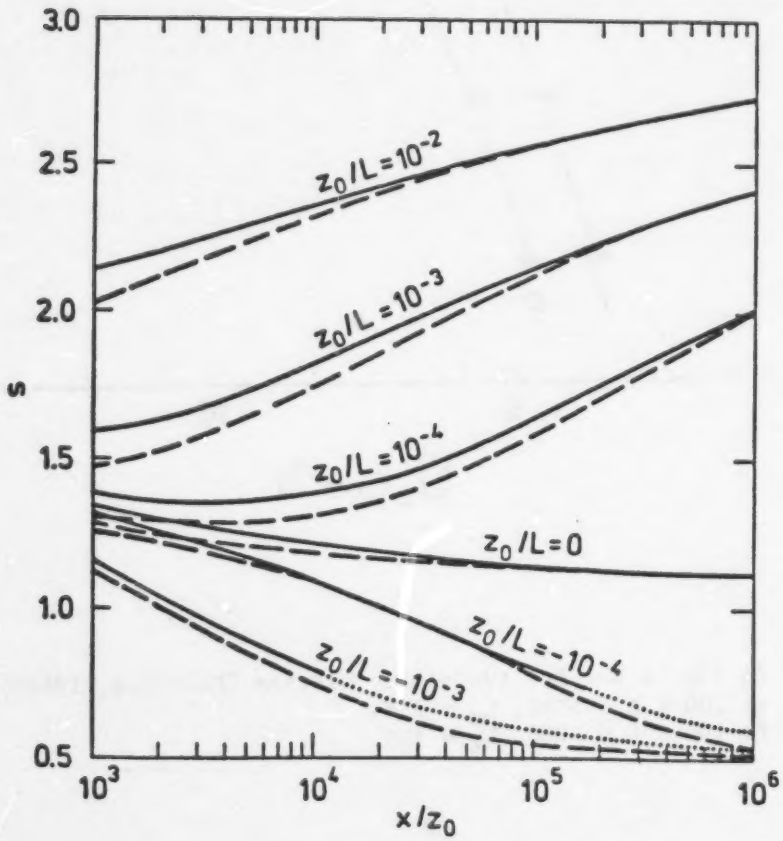

Fig. 6 The shape parameter s of Eq. (8) as a function of downwind distance and stability. The dashed lines show analytical approximations suggested in Gryning et al. (1983). Portions of the curves in which $\bar{z}/L < -1$ are dotted.

obtained. This profile relates the wind speed U_2 at level z_2 to the wind speed U_1 at level z_1 by

$$U_2 = U_1 \frac{\ln(\frac{z_2}{z_o}) - \psi_M(\frac{z_2}{L})}{\ln(\frac{z_1}{z_o}) - \psi_M(\frac{z_1}{L})} . \qquad (5)$$

In Eq. (5) z_o is an representative roughness length for the surroundings as discussed by e.g. Davenport (1960) and Wieringa (1980,1981). Table 1 gives z_o in terms of a terrain classification. Further ψ_M is a stability dependent function, which is related to ϕ_m. Paulson (1970) gives ψ_M for m = 1/4 in (4a), while Holtslag (1984b) gives ψ_M for m = 1/3 in (4a). For stable conditions $\psi_M = -5\ z/L$, which results with (5) in the log-linear profile (Hicks, 1976).

Fig. 4 shows the results of Eq. (5) with the two stability functions for ψ_M in unstable conditions, compared with Cabauw data (Holtslag, 1984b). Here the observed 10 m windspeed is used for U_1 and further z_o = 0.2 m (Wieringa, 1980). The observed data are averaged and indicated with dots and error bars of $\pm\ \sigma/\sqrt{n}$, where σ is the rms deviation of observed wind speeds in a given stability class defined by $L_m = \overline{1/(1/L)}$ (n is number of observations). It is seen that for both ψ_M functions the agreement is good up to 200 m, e.g. $-z/L_m \approx 7$ for the most unstable class. Fig. 5 shows Eq. (5) for stable conditions at Cabauw. It is seen that the agreement is good up to $z/L_m \approx 1$. Above the latter height deviations occur, which are discussed in the section for the stable ABL.

In the SL vertical diffusion from a ground level source can be described with the two-dimensional diffusion equation

$$U(z) \frac{\partial \chi_y(x,z)}{\partial x} = \frac{\partial}{\partial z} \{ K(z) \frac{\partial \chi_y(x,z)}{\partial z} \}. \qquad (6)$$

Here $\chi_y(x,z)$ is the crosswind integrated concentration and K(z) is the vertical eddy diffusivity for matter (x is travel distance, z is height). In practice K(z) can be expressed as the diffusivity for heat (Nieuwstadt and Van Ulden, 1978; Gryning et al., 1983)

$$K(z) = \frac{k\ u_*\ z}{\phi_h(\frac{z}{L})} . \qquad (7)$$

Eq. (7) is based on Monin-Obukhov similarity theory, where ϕ_h is the dimensionless temperature gradient. For unstable conditions $\phi_h = \phi_m^2$ and for stable conditions $\phi_h = \phi_m$ (Dyer, 1974).

An approximate solution to (6) can be written as (Van Ulden, 1978; Pasquill and Smith, 1983)

$$\frac{\chi_y(x,z)}{Q} = \frac{A}{\bar{z}\ \bar{U}} \exp(-(B\ \frac{z}{\bar{z}})^s) , \qquad (8)$$

where Q is source strength, A and B are functions of the shape parameter s, $\bar{z}$ is the mean height of the plume and $\bar{U}$ is mean wind speed across the plume. Fig. 6 shows calculated values of s with (6) and (8) when no flux of matter to the surface (deposition) is taken into account (solid line)

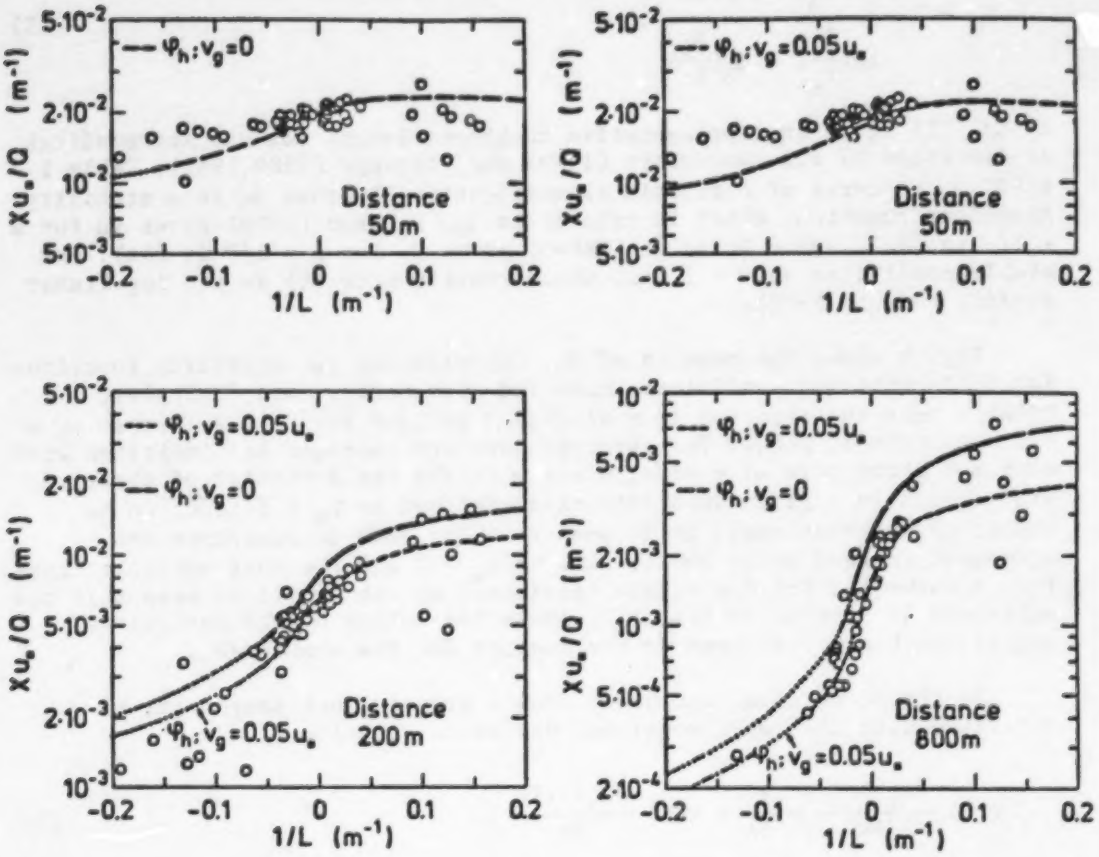

Fig. 7 Predicted and measured crosswind-integrated concentrations at the surface $\chi_y u_*/Q$ (m^{-1}) at 50, 200 and 800 m downwind from the source as a function of stability 1/L (m^{-1}) for project Prairie Grass (Gryning et al., 1983). Portions of the curves in which $\overline{z}/L < -1$ are dotted.

(Gryning et al., 1983). It is seen that s is a function of downwind distance and stability. The dashed line in Fig. 6 is an analytical approximation to the results for s given by Gryning et al. (1983). Van Ulden (1978) gives an approximation for $\overline{U}$ and $\overline{z}$ when s is fixed on s = 1.5.

Fig. 7 shows the numerical solution of (6) for $\chi_y u_*/Q$ (m^{-1}) at the surface as a function of 1/L (m^{-1}) (Gryning et al., 1983). In the figure two cases are distinguished. In the first case deposition is taken zero ($v_g = 0$), while in the second case the lower boundary condition on (6) is given by $K\ \partial\chi_y/\partial z = v_g\ \chi_y$ for $z = z_o$. Here v_g is the deposition velocity. Those portions of the curves in which $\overline{z}/L < -1$ (above the unstable SL) are dotted. The calculations are compared with data of the Prairie grass experiment with an averaging time of 10 min. (Barad, 1958). Overall Fig. 7 shows that the agreement of (6) with these observations is good. On the distance of 800 m incorporation of $V_g = 0.05\ u_*$ improves the agreement with the observations. For a more detailed discussion on the application of (6) we refer to Nieuwstadt and Van Ulden (1978), Van Ulden (1978), Gryning et al. (1983) and Pasquill and Smith (1983).

THE FREE CONVECTION AND MIXED LAYER

The free convection layer (FCL) and the mixed layer (ML) occur above the surface layer (SL) typically for $-h/L > 10$ (Fig. 3). However, in Fig. 3 it is indicated that a value of $-h/L \approx 5$ may be sufficient to drive the ABL in the convective state (Tennekes, 1982). In that case the corresponding top of the SL is $-z/L = 0.5$.

In the FCL and ML the momentum flux τ_o is no longer an important scaling parameter, because wind shear plays no dominant role anymore (Panofsky, 1978). In the ML the relevant scales are therefore $\overline{w\theta_o}$ and h. From these quantities and g/T we can form a velocity scale (Deardorff, 1970; Tennekes, 1970)

$$w_* = \left(\frac{g}{T}\, h\, \overline{w\theta_o}\right)^{1/3} . \qquad (9)$$

It turns out that in the ML turbulence and dispersion scale with w_* and h (Deardorff and Willis, 1975; Lamb, 1982). This means that dimensionless groups formed with w_* and h become universal functions. We can define dimensionless distance $X = w_*\, t/h$, where t is travel time $t = x/U$; dimensionless cross-wind integrated concentration at the surface $C_y = \chi_y(x,0)h\ U/Q$ and dimensionless source height $Z_s = z_s/h$, where z_s is actual source height.

For an elevated source emitting neutrally buoyant material into the ML, it is found that close to the source the line of maximum concentration descends until it reaches ground level (Willis and Deardorff, 1978; Lamb, 1982). This behaviour can be explained by the statistical properties of the turbulence in convective conditions. Throughout most of the layer, except close to ground-level and the inversion base, downdrafts occupy a greater area than updrafts. A particle which randomly is released therefore has a higher probability to descend than to ascend. The consequence is the descent of the plume axis (Baerentsen and Berkowicz, 1984; De Baas et al., 1985).

On the other hand material released near the ground is found to be rising. The reason for this is the fact that the turbulent structure is strongly inhomogeneous near the surface. In particular the condition $d\sigma_w^2/dz > 0$ seems to be crucial (Van Dop et al., 1985; De Baas et al.,

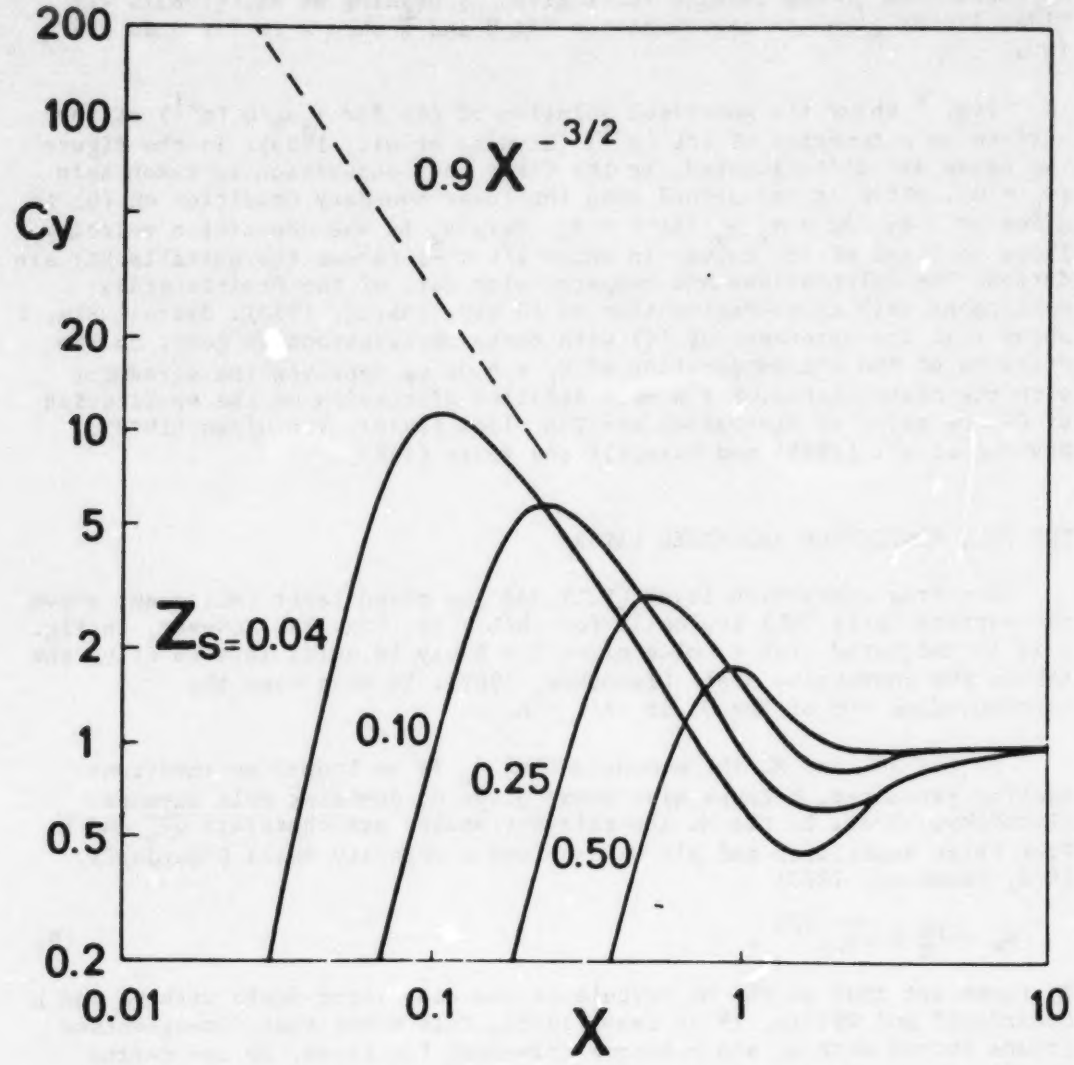

Fig. 8 Plot of dimensionless crosswind-integrated concentration at the surface C_y as given by Eq. (10) as a function of dimensionless distance X and for indicated values of dimensionless source height Z_s. The dashed line is Eq. (12).

1985). Here σ_w is the standard deviation of vertical velocity.

Summarizing we may say that characteristics of dispersion in convective conditions are dominated by the asymmetric structure of turbulent updrafts and downdrafts. Therefore the dispersion depends on source height. Wyngaard and Brost (1984) have called this property "top-down and bottom-up diffusion". In a later paper Wyngaard (1984) discusses some additional results along these lines. However, this approach is still limited to steady-state conditions and has not yet been applied to the transient behaviour of point sources.

The behaviour of contaminants discussed above is observed in watertank experiments (Willis and Deardorff, 1978) and numerical experiments (Lamb, 1982). The results are confirmed by observations of Poreh and Cermak (1984) in a wind tunnel and by field experiments (Briggs, 1983). On basis of these results Briggs provided an empirical relation between C_y at the surface, X and Z_s which reads (see Irwin et al., 1985)

$$C_y = \frac{0.9\, X^{9/2}\, Z_s^{-11/2}}{[Z_s^{-3/4} + 0.4\, X^{9/2}\, Z_s^{-9/2}]^{4/3}} + \frac{1}{[1 + 3\, X^{-3/2}\, Z_s^{1/2} + 50\, X^{-9/2}]} \,. \qquad (10)$$

This expression can according to Briggs be applied for non-buoyant releases when $0.04 \leq Z_s \leq 1$.

Fig. 8 shows (10) for four values of Z_s. For all source heights the plume is well-mixed in the ABL beyond X = 5. Below the latter value Z_s is important for C_y, as is indicated by the peak surface concentrations and the subunity values of C_y near X = 2. The latter is associated with an elevated concentration maximum (Deardorff and Willis, 1975; Briggs, 1983), which can be explained by the inhomogeneous turbulent structure near the surface.

In Fig. 9 we have compared the results of (10) with independent dispersion data of non-buoyant elevated releases at Copenhagen (Gryning and Lyck, 1984). The averaging time is 20 min. Included are the results for two experiments for which ML scaling applies at three distances (triangles and crosses). Here the ML is defined as $-h/L > 10$ and $z/h > 0.1$ (Fig. 3). Overall, it is seen in Fig. 9 that the model estimates are lower than the observations and that the deviations increase with X. The performance for the run of July 6, 1979 (triangles) is good, but for the run of October 19, 1978 (crosses) the agreement is less for $X > 0.7$. For the latter run assumption of $C_y = 1$ for $X > 0.7$ would lead to a better agreement with the data, which means that the dip in Fig. 10 is not reproduced by this run. These findings illustrate the uncertainty in estimating dispersion in a convective boundary layer (Venkatram, 1984).

Below the mixed layer (ML) the free convection layer (FCL) exists as a matching region between the surface layer (SL) and the ML (see Fig. 3). In this layer, therefore, τ_o is no longer important and h is not yet. The only relevant length scale is given by z (Panofsky, 1978). Therefore in the FCL the matching conditions imply that equations such as (10) can no longer depend on h. Instead we can make dimensionless combinations of χ_y/Q with U, z, x, z_s and a velocity scale for the turbulence (Tennekes, 1970)

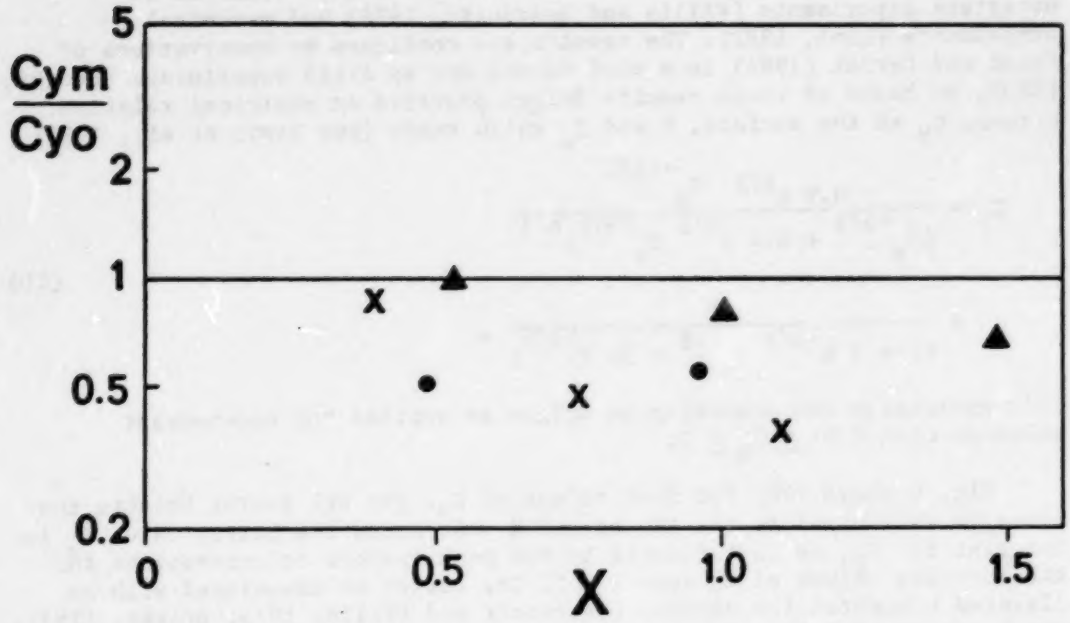

Fig. 9 Comparison of the ratio between model values of C_y at the surface and observed values of C_y as a function of dimensionless distance X. Model values (C_{ym}) are obtained from Eq. (10). Observed values (C_{yo}) are from the Copenhagen experiment on three days:
Triangles: run July 6, 1979; $Z_s = 0.14$, $-h/L = 11.3$.
Crosses: run October 19, 1978; $Z_s = 0.103$, $-h/L = 10.4$.
Dots: run September 20, 1978; $Z_s = 0.058$, $-h/L = 43$.

$$w_f = \left(\frac{g}{T} z \overline{w\theta_o}\right)^{1/3} . \qquad (11)$$

At present such relations do not exist for elevated releases. However, Eq. (10) may be used already for $Z_s > 0.04$, according to Briggs analysis. Fig. 9 shows the results of (10) for a Copenhagen run which is in the FCL (dots). The observations are, however, a factor two larger than the model estimates. Again assumption of $C_y = 1$ would lead to better results for this run as well.

For surface releases Nieuwstadt (1980) pointed out that in the FCL

$$C_y = 0.9\ X^{-3/2}, \qquad (12)$$

fits the Prairie grass surface concentrations very well for $0.03 < X < 0.23$. With the definitions of C_y, X and w_* it can be shown that in (12) the dependence on h disappears. In Fig. 8 Eq. (12) is indicated by a dashed line. The FCL predictions by (12) were compared by Holtslag (1984a) with the SL predictions of (8) with $s = 1.5$ (Van Ulden, 1978). The comparison was done on the Priarie grass data. At $x = 50$ m it was seen that the SL model performed better than the FCL model, confirming that at $x = 50$ m the plume was still within the SL. At $x = 200$ m and $x = 800$ m the predictions of the two models were found to be similar. This suggests that both scaling principles can be used to these distances. The borders between the scaling regions as depicted in Fig. 3, therefore, only give an order of magnitude for the separation of the scaling regions.

In Fig. 4 we have seen that the SL wind profile (5) showed good agreement with data up to even $-z/L \approx 7$. This means that (5) can be used also in the FCL. In the ML it is best to assume a uniform profile of U with the value obtained with (5) at $z = 0.1$ h (Garratt et al., 1982).

THE STABLE ABL

As discussed before, the mean height of the stable turbulent ABL is typically an order of magnitude smaller than that under unstable conditions. Due to the stable stratification in the ABL the turbulence is suppressed, which leads to smaller eddies in the stable ABL. As a consequence the structure of the stable ABL is completely different from that in the unstable ABL. Nevertheless, the turbulence in the lower layer of the stable ABL scales with z, τ_o and $\overline{w\theta_o}$ as in the unstable ABL. Above this SL the turbulence scales with z and the local values of the fluxes τ and $\overline{w\theta}$. This is known as local scaling, a similarity approach recently developed by Nieuwstadt (1984a).

With the local fluxes a local Obukhov length Λ can be formed, which reads

$$\Lambda = -\frac{\tau^{3/2}}{k \frac{g}{T} \overline{w\theta}} . \qquad (13)$$

In the local scaling region only two length scales arise, z and Λ. With these the turbulent quantities can be described as a function of z/Λ (Nieuwstadt, 1984a). For larger values of z/Λ local scaling reduces to z-less scaling, meaning that z is no longer an important length scale for the turbulence. The background is that vertical motion is inhibited due to the stable stratification and that turbulent eddies no longer feel the presence of the surface. Consequently, z-less means that dimensionless

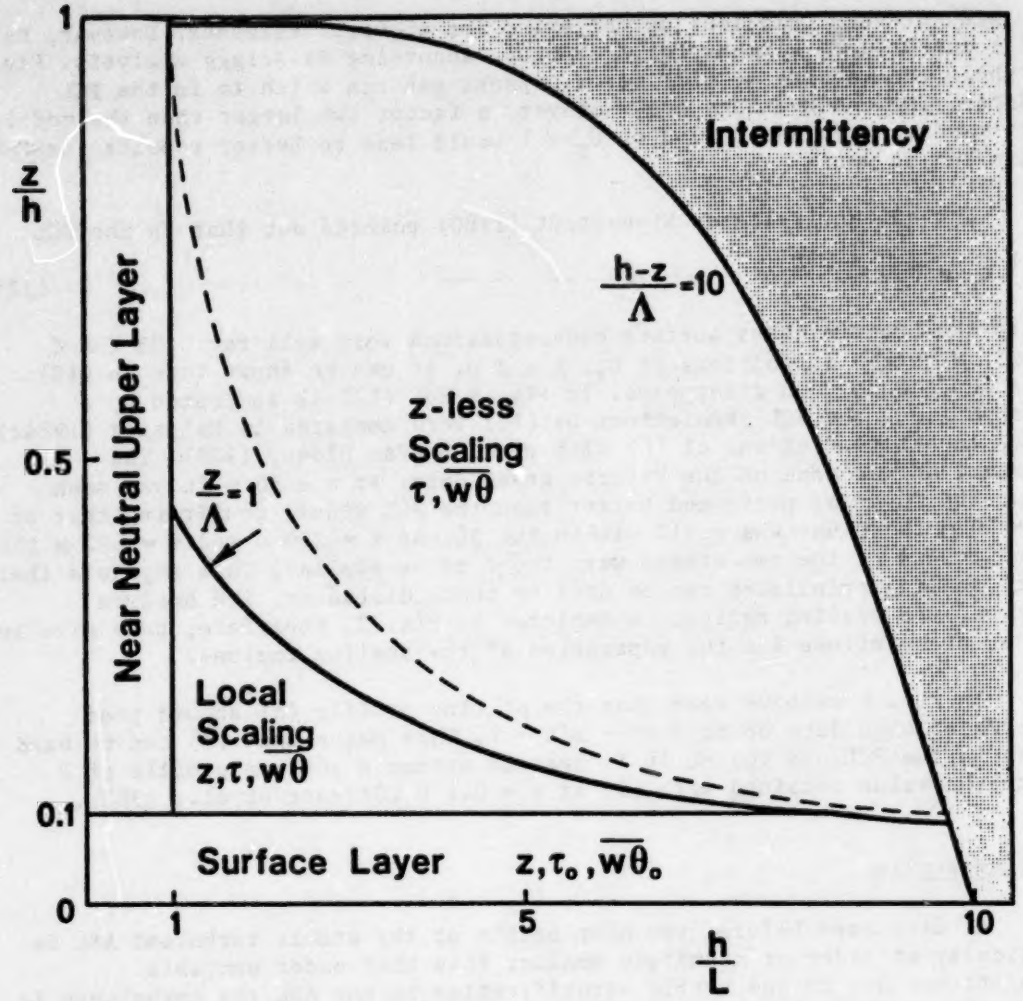

Fig. 10 Definition of scaling regions in the stable ABL (L > 0) with indicated basic scaling parameters for the turbulence (Holtslag and Nieuwstadt, 1985). The dashed line is given by z/L = 1. Further explanation is in the text.

quantities approach a constant value for large z/Λ. For instance, the flux Richardson number Ri_f approaches a constant value of $Ri_f = 0.2$. This value is already achieved by the Cabauw data at $z/\Lambda = 1$ (Nieuwstadt, 1984a).

For the application of local scaling and z-less scaling, profiles of τ and $\overline{w\theta}$ are needed through the whole turbulent ABL. These profiles can in principle be obtained from the equations which describe the time development of the stable ABL. This means that in general the vertical profiles are a function of time and are dependent on the time history of boundary layer development. So we can not expect a generally valid similarity profile to describe vertical profiles. However, in the special circumstance of horizontally homogeneous and steady conditions when the cooling rate of the ABL is constant, Nieuwstadt (1984a) was able to derive explicit profiles. This was done by applying the closure hypotheses $Ri = Ri_f = 0.2$ on the equations for the stable ABL, where Ri is the Richardson number.

The steady state profiles for τ and $\overline{w\theta}$ read

$$\tau/\tau_o = (1 - z/h)^{3/2}, \tag{14}$$

with $\tau_o = u_*^2$ and

$$\overline{w\theta}/\overline{w\theta}_o = (1 - z/h). \tag{15}$$

With (2), (13), (14) and (15) we obtain for Λ:

$$\Lambda/L = (1 - z/h)^{5/4}. \tag{16}$$

The solution for the steady state ABL equations also lead to an expression for the turbulent boundary layer height

$$h = c\left(\frac{u_*}{f} L\right)^{\frac{1}{2}}, \tag{17}$$

equivalent with the expression found by Zilitinkevich (1972). The constant c is given by $c = (\sqrt{3}\, k\, Ri_f)^{\frac{1}{2}} = 0.4$. Remember that h may be a fraction of the surface inversion depth caused by radiation (Wetzel, 1982).

It was shown by Nieuwstadt (1984a) that (14) and (15) are in good agreement with observations in Cabauw, even in non-steady conditions. The evolution of h during nighttime should be described in terms of a prognostic equation as discussed by Nieuwstadt (1984b). Nevertheless, in the latter study Nieuwstadt showed that Eq. (17) can be used as a first approximation with generally good results. In Fig. 2 we have indicated the estimate of h with (17) during nighttime. It is seen that indeed the agreement with observations is good. These results are probably only applicable above rather homogeneous terrain without a slope (as in Cabauw).

In Fig. 10 taken from Holtslag and Nieuwstadt (1985), the above scaling is summarized. Here conditions with $h/L < 1$ are considered near neutral (Nieuwstadt, 1984b). The SL is taken up to $z/h = 0.1$ in agreement with the linear profile (15) and our discussion for the SL. The local scaling region is defined up to $z/\Lambda = 1$, because z-less scaling already applies for $z/\Lambda \gtrsim 1$ as indicated by Nieuwstadt's results. The intermittency region of Fig. 10 indicates a region, in which the turbulence is not continuous in time and space anymore. This means that the turbulence is confined to isolated patches, which develop and disappear as a function of time. Intermittency occurs at low wind speed or large stability. In fact no fully turbulent stable ABL's are observed

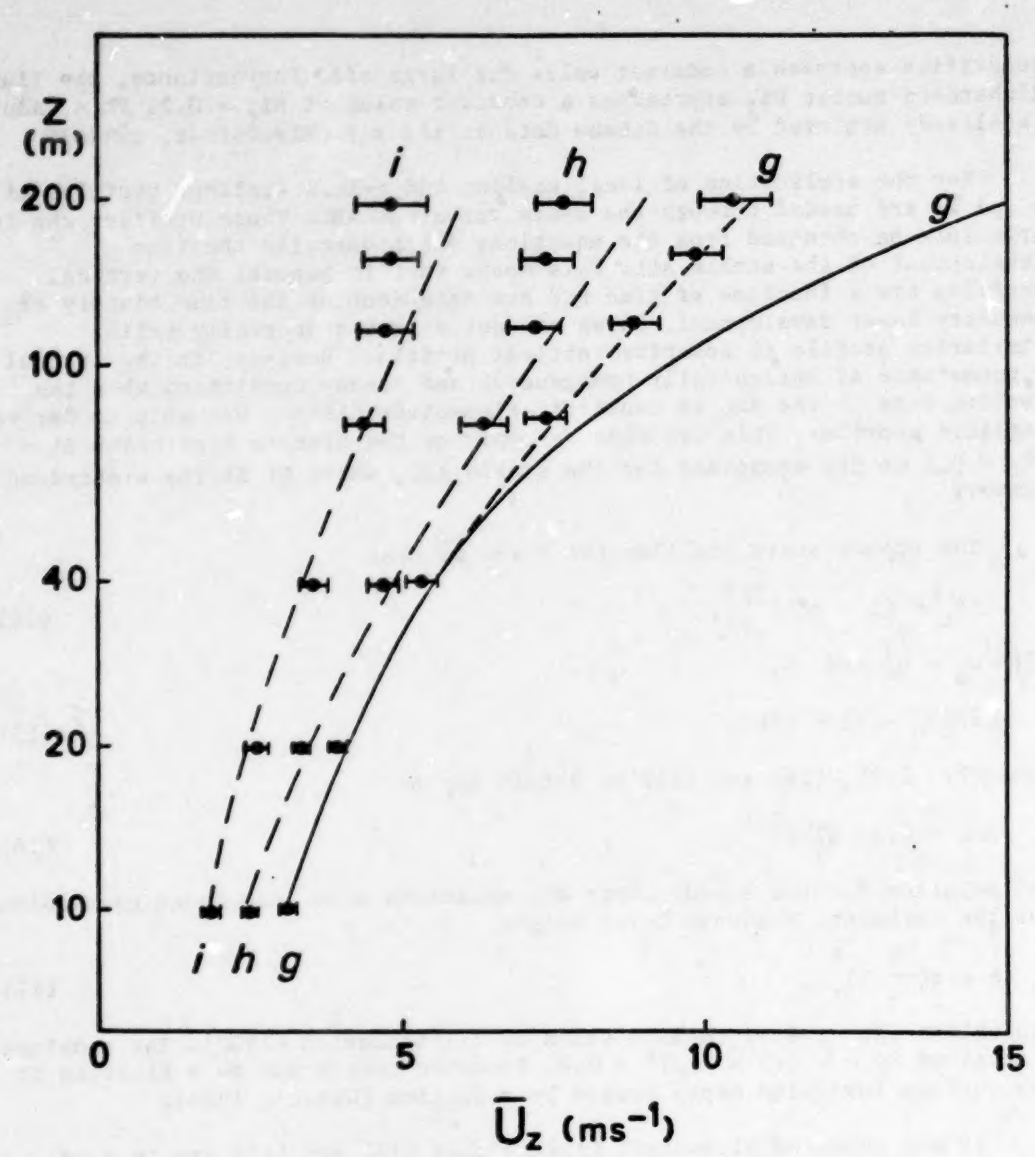

Fig. 11 As Figs. 4 and 5 for three very stable classes (Holtslag, 1984b):
g: $40 < L \leq 100$, $L_m = 60$.
h: $10 < L \leq 40$, $L_m = 20$.
i: $0 \leq L \leq 10$, $L_m = 9$.

beyond $h/L \approx 10$. For larger stability the present similarity theories break down. No satisfactory theory for the intermittent stable ABL has been developed at present. For $h/L < 10$ turbulence is also intermittent near the top of the stable ABL. This is connected to fluctuations in the instantaneous value of h. In Fig. 10 the intermittency region is estimated as $(h-z)/\Lambda > 10$. The border line $(h-z)/\Lambda = 10$ is continuous with $h/L = 10$ at the surface. This line shows that with increasing stability the region of fully turbulence decreases. Comparison of Fig. 10 with Fig. 3 shows clearly the difference in the turbulent structure between the unstable and stable ABL.

In Fig. 10 also the line $z/L = 1$ is indicated (dashed). Comparison of this line with the line $z/\Lambda = 1$ shows that in a large part of the stable ABL $\Lambda = L$ (see also Eq. (16)). This means that a formulation of z/Λ as required by local scaling is almost equivalent to the surface layer similarity scaling in terms of z/L. We can use this property for the estimation of wind profiles above the SL. Fig. 11 shows three very stable wind profiles at Cabauw (Holtslag, 1984b). As in Figs. 4 and 5 the data is averaged in well defined stability classes of Obukhov length L. For class g the usual SL theory predicts the indicated full line in agreement with the data up to ~ 60 m. For class g the average values for L and h are estimated by Holtslag as ~ 60 m and ~ 160 m, respectively. As in Fig. 5 we obtain good agreement up to $z/L \approx 1$, which in class g corresponds to $z/h \approx 0.4$ far above the SL! Above the height of ~ 60 m large deviations occur between the data and the usual SL theory. For the other classes of Fig. 11 no agreement is found at all (not shown here).

In Fig. 11 the dashed lines are in good agreement with the data up to the estimate of h with Eq. (17) with a minimum of ~ 100 m (Holtslag, 1984b). These lines were calculated with Eq. (5), using a suitable expression for ψ_M given by (Holtslag, 1984b)

$$\psi_M = -7 \ln\left(\frac{z}{L}\right) - \frac{4.25}{(z/L)} + \frac{0.5}{(z/L)^2} - 0.852 \quad . \tag{18}$$

This expression for ψ_M is continuous with the usual SL expression $\psi_M = -5\, z/L$ for $z/L = 0.5$. Eq. (18) is based on findings by Hicks (1976) for very stable Wangara data. For the turning of wind with height and stability and further discussion of the data in Fig. 11 we refer to Holtslag (1984b) and Van Ulden and Holtslag (1985).

For elevated releases in the stable ABL Nieuwstadt (1984b) used a K-model, where the formulation for K was consistent with the local z-less scaling approach. Another possibility for the description of vertical diffusion in the stable ABL is given by statistical theory. In that case the vertical dispersion coefficient σ_z is written as (Draxler, 1976)

$$\sigma_z = \sigma_w\, t\, f_z(t/T_e), \tag{19}$$

where t is travel time of the plume ($t = x/U$), σ_w is standard deviation of vertical velocity, T_e is time scale for diffusion and f_z is given by

$$f_z = \left[1 + 0.9\left(\frac{t}{T_e}\right)^{\frac{1}{2}}\right]^{-1} \quad . \tag{20}$$

For stable conditions T_e can be approximated by $T_e = 50$ s for an averaging time of 30 min., typically (Irwin, 1983). However, statistical theory assumes homogeneous turbulence, which is certainly not satisfied in the stable ABL. This is due to the strong variation of profiles with height. Therefore, (19) and (20) should be applied with caution. Nevertheless, Venkatram et al. (1984) obtained good agreement for σ_z with observations, using (19) and a slightly different formulation for f_z. The time scale in

their form of f_z is estimated as $T_L = \ell/\sigma_w$, where ℓ is a "mixing length scale" (Hunt, 1982). The value of ℓ is obtained by Venkatram et al., with the use of local scaling arguments. It can be shown that for very stable conditions ℓ is proportional to Λ.

For the application of (19) a value for σ_w is needed, preferably at source height z_s. When no measurements are available we can estimate σ_w from local z-less scaling. In that case $\sigma_w/\tau^{\frac{1}{4}} = c$, where c is a constant. This results with (14) in (Nieuwstadt, 1984b)

$$\sigma_w/u_* = c(1 - z_s/h)^{3/4}. \qquad (21)$$

Here for c = 1.3 Eq. (21) agrees with the value for σ_w in the surface layer (e.g. Caughey, 1982). With (19)-(21) we can obtain σ_z from u_*, z_s, h, t and T_e. This result can be used in a Gaussian plume model (Pasquill and Smith, 1983). Comparison of these model results with ground level concentrations awaits further examination.

Other modeling attempts of diffusion in the stable ABL have been proposed by Pearson et al. (1983). They describe diffusing in terms of particle movements in a stable environment. They find that under strong stability the vertical plume dimension is limited and the growth is very slow, due to exchange on molecular scales. However, their results were not confirmed by the findings of Venkatram et al. (1984).

LATERAL DISPERSION

In the preceding discussion the structure of the turbulence and its implication on the vertical dispersion process was described. Scaling lengths of the flow were imposed by the dimensions of the flow (e.g. depth, distance from ground). Because such restrictions do not exist in the lateral direction there are no obvious length scales. However, as the size of the eddies are controlled by the vertical scaling lengths, these put some constraints on the flow in the lateral direction as well (Deardorff and Willis, 1975).

Experiments show (e.g. Sivertsen, 1978; Van Duuren and Nieuwstadt, 1980; Gryning, 1981) that the lateral distribution of a plume resembles a Gaussian distribution. Deviations from a Gaussian distribution exists for the individual cases, but ensemble averaged concentration distributions are Gaussian in appearance. Therefore, knowing the cross-wind integrated concentration at the surface $\chi_y(x,o)$, we can calculate the surface concentration from

$$\chi(x,y,o) = \frac{\chi_y(x,o)}{(2\pi)^{\frac{1}{2}}\sigma_y} \exp\{-\frac{y^2}{2\sigma_y^2}\}, \qquad (22)$$

where y is crosswind direction and σ_y is the crosswind spread of the plume.

The cross-wind plume spread is best calculated by use of Taylors statistical theory for plume dispersion (Draxler, 1976; Pasquill and Smith, 1983; Gryning and Lyck, 1984). With this theory the width of a plume σ_y, is related to the standard deviation of wind σ_v under homogeneous, stationary conditions. Because for the lateral dispersion no rigid restrictions exist, the assumption of homogeneity is fairly well obeyed. Taylors formula can be expressed as (Pasquill and Smith, 1983)

$$\sigma_y = \sigma_v \, t \, f_y(t/T_y), \tag{23}$$

where the dimensionless spread, $\sigma_y/(\sigma_v t)$ is suggested to be function of the dimensionless travel time, t/T_y. T_y is a time scale for the lateral dispersion process. The assumptions that are imposed in order to derive Eq. (23) are discussed in Gryning (1981) and Pasquill and Smith (1983).

A comparison of several empirical forms of f_y with field data has been performed by Irwin (1983) and by Gryning and Lyck (1984). The general conclusion is that the scheme proposed by Draxler (1976) is best overall. The f_y is for an averaging time of the order of 30 min. to 1 h given by

$$f_y = \frac{1}{1 + 0.9(t/T_y)^{0.5}} \, . \tag{24}$$

For practical use it is suggested that $T_y = 1000$ s for elevated sources, and $T_y = 300$ s for ground level sources. However, Irwin (1983) obtained good results with $T_y = 1000$ s for both elevated and ground level sources. Eq. (24) performs best if measured values of σ_v at source height z_s are used. The measured values should represent an averaging time that corresponds to the averaging time of the concentration values.

When measurements of σ_v are not available, this parameter can be estimated from

$$(\sigma_v/u_*)^2 = 0.7(-h/L)^{2/3} + 2.7 \frac{(1 - z_s/h)^2}{(1 + 2.8 \, z_s/h)^{2/3}} \tag{25}$$

as given by Højstrup (1982) for unstable conditions, and

$$(\sigma_v/u_*)^2 = 2.7 \frac{(1 - z_s/h)^2}{(1 + 2.8 \, z_s/h)^{2/3}} \tag{26}$$

under stable conditions (Irwin et al., 1985). Eq. (25) is a combination of shear produced variance and buoyancy produced variance. The corresponding averaging time for Eqs. (25)-(26) is roughly ~ 30 min. The expression for σ_v under stable conditions (26) does not include the effect of waves which for longer averaging times (> 10 min) can be very significant; again stressing the importance to measure σ_v.

From Eqs. (23)-(26) σ_y can be obtained as a continuous function of t, T_y, u_*, h/L and z_s/h. For convective conditions the second term in (25) is small. In that case (25) results in $\sigma_v \sim w_*$ using the definitions of L and w_* (Eqs. (2) and (9)). With $T_y \sim h/w_*$ we obtain that σ_y is a function of w_*, h and travel time t ($\approx x/U$). Such expressions for σ_y are discussed by Deardorff and Willis (1975), Nieuwstadt (1980) and Briggs (1983).

The lateral dispersion can be viewed as the combined effect of dispersion induced by turbulence of the air, and by shear in the wind direction. Only the first of these processes has been discussed here. The effect of shear is not believed to be important within the first 5-10 km from the source, except under very stable conditions. Even when the actual conditions are close to ideal, the stochastic nature of the dispersion process will give rise to substantial differences between real (measured) and predicted quantities. For σ_y, the fractional error between measured and modeled values was ~ 15%, for the Copenhagen experiment (Gryning and Lyck, 1984). The corresponding figure for the concentrations was ~ 40%.

SUMMARY AND DISCUSSION

In this paper we have reviewed the characteristic features of the turbulent ABL by means of scaling arguments. To incorporate the proper scaling parameters the discussion was structured in accordance with current concepts for the idealized states (regimes) of the ABL. We limited ourselves to meteorological conditions affecting the dispersion for transport on the order of 10 km or less. Often the distance to the maximum ground-level concentration is within this range.

For several states of the ABL procedures are suggested for the mean wind profile and vertical dispersion for non-buoyant releases. Generally, vertical dispersion is found to be other than Gaussian. Therefore, whenever justified we recommend the use of techniques that characterize directly the cross-wind integrated concentration at the surface (χ_y). For the lateral dispersion a Gaussian distribution is assumed for which the input parameters are suggested. Then with χ_y the ground level concentrations can be calculated.

The basic meteorological parameters needed for the methods discussed, are the surface fluxes of heat and momentum, and the mixing height. For the derivation of the surface fluxes a reliable method is the profile method (Berkowicz and Prahm, 1982; Van Ulden and Holtslag, 1985). This method needs at least a wind speed observation at a single height in the surface layer, mean temperature at two heights in the surface layer and an estimate of the surface roughness length z_o (see table 1).

Often, temperature is observed at only one height near the surface, as is the case on routine weather stations. In that case the surface fluxes can be parameterized in terms of a single wind speed observation, total cloud cover, air temperature and solar elevation during daytime (Holtslag and Van Ulden, 1983). During nighttime the surface fluxes can be parameterized in terms of total cloud cover and a single wind speed observation (Venkatram, 1980; Van Ulden and Holtslag, 1983, 1985).

A prognostic mixed-layer model can be used during daytime for the derivation of mixing height h (Driedonks, 1982; Driedonks and Tennekes, 1984). In that case at least a temperature sounding is needed from a radiosonde ascent and the actual (parameterized) surface fluxes. During nighttime h should be described in terms of a prognostic equation as well (Nieuwstadt, 1984b). However, Nieuwstadt also showed that Eq. (17) can be used as a first approximation with generally good results.

Wilczak and Phillips (1984) showed that the meteorological parameters during daytime can be estimated with an accuracy of 10 to 30% from the type of methods discussed above. A review on the derivation of meteorological parameters needed for air pollution dispersion models is given in Van Ulden and Holtslag (1985). A minimum local measurement program is discussed by Irwin et al. (1985). Hoffnagle et al. (1981) discuss instrument requirements for this purpose. Additional special measurements of turbulent quantities can be utilized by the methods described in this paper.

To conclude, we have shown that the transport and dispersion of air pollution can be related directly to the turbulent state of the ABL. The methods outlined should be tested against independent data. With the test results future research can focus on estimation methods having the best agreement. Improvements may be found in a continuous description of vertical dispersion between the different scaling regions of the ABL. Moreover, methods for the estimation of plume rise in the different

regions of the ABL have to be included (Deardorff and Willis, 1984).

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